Columbia University
in the City of New York

LAMONT GEOLOGICAL OBSERVATORY
PALISADES, NEW YORK

THE UNCONSOLIDATED SEDIMENTS

by

John I. Ewing and John E. Nafe

Technical Report No. 3
CU-4-61 NObsr 85077 Geology

June 1961

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T H E   U N C O N S O L I D A T E D   S E D I M E N T S

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CONTENTS

I INTRODUCTION

In this discussion the unconsolidated sediments are taken to be those in which the speed of compressional waves is less than about 4 km/sec. The upper limit is entirely arbitrary and has been chosen to include all the material with velocities lower than that of "Layer 2," which is discussed by Raitt (1956). Layers in which the velocity is between 2 and 4 km/sec have been described by various investigators as semi-consolidated sediments. Other authors have designated layers as unconsolidated A, B, ... , semi-consolidated A, B, ... , and consolidated A, B, ... In almost every case the designation has been made solely on the basis of seismic wave velocity, since identification according to sediment type, age, or stratigraphic position has not been possible except in certain shallow water areas where wells have been drilled nearby.

Most of the marine seismic measurements reported to date have been made in the Atlantic and Pacific oceans and in the shallower inter-continental seas. In all areas investigated, the thickness of the unconsolidated sediments varies widely from zero on some of the topographic highs to several kilometers in some of the deeps and in some continental rise areas. The relationship of sediment thickness and topographic setting has become increasingly more understandable, and even predictable, in recent years since Kuenen introduced the concept of transport of sediments by density or turbidity currents. This concept has been thoroughly verified by sediment coring and seismic measurements. It is now well known that even the coarser sediments derived from continental erosion can be carried great distances from land, flowing out into the ocean basins and being empounded in the deepest accessible areas. It is also known from underwater photography, coring, dredging, and seismic measurements that many topographic highs are devoid of even pelagic sediments, presumably being kept clean by deep currents or by slumping.

In addition to the variation of sediment thickness with topographic setting, there is a marked difference in average thickness of sediments in the Atlantic Ocean and in the Pacific. The Atlantic has an average of more than 1 km (Ewing and Ewing, 1959), compared with less than 1/2 km in the Pacific (Raitt, 1956). This difference is generally attributed to the facts that the Atlantic is much smaller and that more large, sediment-bearing rivers flow into it.

The thickness of sediment cannot always be measured exactly by seismic methods, owing to the difficulty of measuring the velocity, particularly in the uppermost part of the section in the deep ocean. Refracted waves give accurate measurements of the higher velocity layers; but because of various factors such as poor velocity or acoustic impedance contrast between the water and upper sediments, or between two sediment layers, clear reflected or refracted waves from these layers sometimes are not observed. Figure 1 shows the time-distance graph and structure section for a profile recorded in the South Atlantic in an area of thick sedimentary cover. This is typical of the seismic data from which sedimentary velocities and thicknesses are computed. The R_I , R_{II} , and R_{III} curves are determined by reflected waves from the 1st, 2nd, and 4th sub-surface interfaces respectively. The lines G_3 and G_4 are determined by refracted arrivals associated with the tops of layers 4 and 5. Note that no G_1 line (corresponding to refracted arrivals along the sea floor) is observed. This is typical of practically all, if not all, deep ocean areas. The significance of the absence of such arrivals is that in most places the seismic wave velocity in the upper part of the deep ocean sediments must be equal to or less than that in the water at the sea floor. The fact that reflected arrivals are received from this interface indicates that the acoustic impedance, ρC (density x velocity), is different above and below the interface. The water and sediment mixture must be more dense than the water alone, hence the following relationships can be written as typical of the water-sediment interface

$$C_1 \geqq C_2 \qquad \rho_1 C_1 \neq \rho_2 C_2 \qquad \rho_1 < \rho_2 \tag{1}$$

II EVIDENCE FOR GRADIENTS AND LOW-VELOCITY SEDIMENTS

The evidence for the velocity structure shown by the velocity vs. depth curve in Figure 1 is derived from both refracted and reflected waves. In a time-distance graph, there is theoretically a reflection curve for each interface and a refraction line tangent to it, the inverse slope of which gives the velocity in the material below. If there is a velocity discontinuity at each interface and if velocity does not vary with depth in any layer, the time-distance graph will consist of straight line segments (for refracted arrivals) and hyperbolic curves (for reflections). If the velocity increases in each successively deeper layer R_N will cross R_{N-1}. In the case shown by Figure 1, the average velocity $\bar{C}_2$ in layer 2 is higher than that in layer 1, but R_{II} does not cross R_I. This indicates that the longer range arrivals on the R_{II} curve were not reflected from the 2nd interface, as were those at shorter range, but were bent upward by a velocity gradient before penetrating to the depth of the interface. This behavior of reflected arrivals is typical of all ocean basin areas where detailed seismic studies have been reported (Hill, 1952), (Officer, 1955), (Katz and Ewing, 1956), (Nafe and Drake, 1957), and (Ewing and Ewing, 1959). It is convincing evidence that appreciable velocity gradients exist in the deep ocean sediments. Evidence for velocity gradients is found also in refraction data and has been summarized by Nafe and Drake (1957). Refracted arrivals from various depths in the sediments taken from a number of profiles throughout the Atlantic Ocean show a systematic increase of velocity with depth. The average value of gradient is in agreement with that computed from reflection data.

As mentioned before, the fact that refracted arrivals are not received from the water-sediment interface is interpreted to mean that either there is no velocity discontinuity across the interface or the velocity below is lower than that above. We have evidence for both cases. In certain areas the first sub-bottom reflection curve R_{II} is observed to approach a line parallel to and above R_I on the time-distance graph. This can occur only if the velocity in the upper sediments is lower than water velocity. In other areas the R_{II} curve approaches the R_I curve, indicating no velocity discontinuity across the water-sediment interface.

Further evidence for low-velocity sediments has been shown by Officer (1955) and Katz and Ewing (1956) from the study of guided waves traveling in the bottom just beneath the water-sediment interface. These waves have been recorded in many areas of the Atlantic.

They appear at a constant frequency at ranges beyond that of the water-borne refracted wave which grazes the sea floor. To account for the constant frequency and travel time of these arrivals, it is required that they be coupled waves, excited by the grazing ray in the water, and that they travel in a wave guide in the sediments. The wave guide might be a low-velocity layer, as described by Katz and Ewing, or it might be bounded above by a discontinuity and below by a velocity gradient. In either case the coupling would require that the upper sediments have a lower velocity than the bottom water.

The evidence cited above indicates two significant features of deep ocean sediments: (1) they have appreciable velocity gradients, particularly in the upper few hundreds of meters, and (2) the velocity in the uppermost part of the sedimentary column is equal to or lower than that in the bottom water in most places. Direct measurement of sound velocities in sediments (Laughton, 1954), (Hamilton, 1956), Hamilton, et al, 1956), (Shumway, 1956), (Sutton, Berckhemer, and Nafe, 1957) has shown velocities generally ranging between 1.45 and 1.80 km/sec. Most values for deep ocean sediments are in the lower range, in good agreement with the seismic evidence. In the intermediate and shallow areas, higher velocities are found by both types of measurement.

III VARIABLE ANGLE REFLECTIONS

The method used to determine the amount of gradient by seismic reflection profiles has been described in detail by Hill (1952), Officer (1955), and Katz and Ewing(1956). It is a process in which a family of theoretical curves is compared with the observational data and the best fit determined. The curves shown in Figure 2 (from Nafe and Ewing, unpublished manuscript) give theoretical values of $R_S - R_I$ vs. range D for three different values of water depth and gradient, where R_S is the travel time of the wave refracted in the sediments, R_I is that of the bottom-reflected wave, and D is that of the direct water wave. These curves were computed for linear velocity gradients, for which the formulae are

$$\begin{aligned} R_S &= R_{IO} \sec \theta_1 + 2/K \left(\sin h^{-1} \cot \theta_2 - \sin h^{-1} \cot \theta_3 \right) \\ R_I &= R_{IO} \sec \theta_4 \\ D &= \frac{C_1}{C_0} \left[R_{IO} \tan \theta_1 + \frac{2\,C_2}{K\,C_1} \left(\frac{\cos \theta_2 - \cos \theta_3}{\sin \theta_2} \right) \right] \end{aligned} \qquad (2)$$

where R_{IO} is the time for the bottom-reflected wave at zero shot-receiver distance, K is the gradient, C_0 is the velocity in the surface channel, C_1 is the mean velocity in the water, and C_2 is the velocity in the upper part of the sediments. The ray paths, with angles $\theta_1 - \theta_4$ indicated are shown in Figure 1. The theoretical curves of Figure 2 are those corresponding to the upper branch R_S of the refraction curve shown in Figure 3. Each curve begins at the "critical range" corresponding to the minimum distance at which a ray can be returned to the surface by refraction under the conditions of velocity gradient and water depth indicated. The R_D branches of the curves, corresponding to the deeper penetrating rays, are not shown. They join the R_S branches in cusps at the critical range and are curved in the opposite direction. For simplicity it was assumed that the water has uniform velocity equal to that in the upper sediments. Hence, in equations (2), $C_0 = C_1 = C_2$, $\theta_1 = \theta_2$, and since we are dealing only with refracted waves, $\Theta_3 = \pi/2$. For the ranges involved, it can be shown that errors introduced by the assumption that $C_0 = C_1$ are small, and as discussed earlier, we have much evidence that velocity in the upper sediments is approximately equal to that in the water.

These same formulae can also be used for the cases in which the upper sediment velocity is either lower than or higher than water velocity. For $C_1 > C_2$, the curves corresponding to those in Figure 2 would be displaced down and to the right. For $C_1 < C_2$, they would be displaced upward and toward the left. In the first case, there is no critical angle for reflection, even at grazing incidence, hence the R_S arrival is always later than R_I . In the second case, R_S will cross R_I . In the case illustrated, i.e. $C_1 = C_2$, R_S joins R_I at the range corresponding to a grazing ray.

On these theoretical curves are plotted squares and triangles representing the observational data from two reflection profiles, A and B, with water depths corresponding to $R_{IO} = 5$ and 7 seconds respectively. The data are plotted for ranges from 0 to 20 seconds, hence both reflected, R_{II}, and refracted, R_S, arrivals are shown. In profile A, the outer points best fit the refraction curve for $R_{IO} = 5$ and $K = 0.45$. At ranges beyond about 9.3 seconds , the observed arrivals are purely refracted. Since 9.3 seconds is the <u>minimum</u> range at which a refracted arrival can be returned to the surface under the conditions $R_{IO} = 5$ seconds and $K = 0.45\ \mathrm{sec}^{-1}$, the arrivals at shorter range must be reflected. The fact that the reflected arrivals join smoothly to the refracted ones indicates that the reflecting interface must lie at approximately the depth penetrated by the critical ray, i.e. the ray at range 9.3 seconds. If the interface had been deeper, the reflection curve would have crossed the $R_S - R_I$ curve above the cusp and joined the $R_D - R_I$ curve as indicated diagrammatically in Figure 4.

If a reflecting interface is shallower than the depth of penetration of the critical ray, the $R_{II} - R_I$ curve approaches and joins the $R_S - R_I$ curve from below as demonstrated by the arrivals of profile B in Figure 2. For this profile $R_{IO} = 7$ seconds and the best fit to the data is with the curve for $K = 1.0\ sec^{-1}$.

Figure 5 shows the results from another profile for which $R_{IO} = 6$ seconds and $K = 0.9\ sec^{-1}$. There are two reflecting interfaces, both shallower than the depth of penetration of the critical ray, as evidenced by the fact that both reflection curves pass below the cusp of the refraction curves for $K = 0.9\ sec^{-1}$. From the fact that both curves indicate approximately the same value of K, it would appear that the upper reflector is a thin layer, above and below which the velocity and velocity gradient are not greatly different. This can be stated only qualitatively because the precision of picking the arrivals in the region where the curves come together is not high.

In the profiles shown in Figure 1, the gradient in layer 2 is found by the methods just described to be approximately $0.5\ sec^{-1}$. There is some evidence that the velocity continues to increase with depth below the 2-3 interface. This comes from the relationship of the R_{II} curves and the G_3 lines. Although nearly so, the observed portions of the refraction lines are not tangent to the reflection curves as would be the case if they are associated with the same interface. Certainly it is possible that two interfaces are present; the upper one producing reflections but no refractions and the lower one producing refractions but no reflections, but this seems an unlikely interpretation. In order for the extension of the refraction lines to be tangent to the R_{II} curve, they would have to curve downward in the manner indicated by the dashed lines, indicating a gradient below the 2-3 interface. The fact that the outer portions of the G_2 lines are observed as straight lines indicates that at some depth below the 2-3 interface, the velocity becomes constant or nearly so.

The 2-3 interface is judged to be a discontinuity with the higher velocity below on the basis of the observation of a critical angle for reflections from it. Figure 6 shows four records of the reflections at ranges of 2.6, 3.5, 4.4, and 5.4 seconds from profile 99. Note the sharp change in character and strength of R_{II} in this relatively short range. The same behavior was noticed in profile 98 and is a common feature of most sub-bottom reflection profiles in deep water areas.

The 4-5 interface is determined by the refracted arrivals corresponding to G_4 and by reflected arrivals R_{III}. Samples of the R_{III} signals are shown in Figure 6 in addition to the R_{II} arrivals. In the case of the 4-5 interface, note that there is perfect agreement between reflected and refracted arrivals (Figure 1) in that the G_4 lines are tangent to the R_{III} curves.

As mentioned earlier, it is not unusual to record strong reflections from within the sedimentary column without receiving the corresponding refracted arrivals. This can be an indication that the reflector is a thin layer, or it could in some cases be explained if the velocity discontinuity is to a lower velocity below. Still another explanation can be that the discontinuity is one of density rather than of velocity, i.e. $C_1 = C_2$; $\rho_1 C_1 \neq \rho_2 C_2$.

The foregoing has been an attempt to describe the methods of obtaining velocity structure in the sediments in deep ocean areas, and some specific results have been shown. Although the cases discussed were confined to comparison with linear gradients, comparisons have been satisfactorily made with parabolic and exponential curves. It is improbable that any one type of gradient is applicable everywhere and, indeed, in most cases the best fit might be obtained with a combination, i.e. linear to a certain depth, and parabolic below that. Previously reported results on gradients measured by seismic techniques have been given by Hill (1952), Officer (1955), Katz and Ewing (1956), and Nafe and Drake (1957) in addition to laboratory measurements on artificially compacted sediments by Laughton (1954). These results and others unreported have shown average gradients, principally from Atlantic Ocean profiles to vary between 0.4 and 2.5 sec^{-1} for the upper 0.3 to 1.0 km of sediments. The variations are undoubtedly due in part to observational or interpretational differences. Some profiles are shot with more appropriate charge sizes or shot spacing than others, and some have much simpler interpretations than others. However, a considerable part of the variations is probably real, indicating differences in sediment types, rates of sedimentation, porosity, grain size, lithification, and in other factors affecting seismic velocity (Sutton, Berckhemer and Nafe, 1957). At present there has not been a sufficient number of determinations to permit the classification of areas by sediment velocity structure except possibly, as suggested by Nafe and Drake (1957), to predict that in areas where calcium carbonate deposition is high, sediment velocities will also tend to be high. Most of the better determinations have been made in abyssal plains or other flat lying areas in order to avoid topographic complications, and results from areas which receive no turbidity current deposits, if available, might show significant differences.

Table I gives values of K from several profiles determined by fitting the theoretical curves for linear gradients. The highest value is that measured by Hill (1952). The lowest was measured on VEMA-15, profile A of Figure 2. The low value of gradient in this case is not unreasonable, because it is the average gradient in approximately 1 km of sediments. The higher values come from profiles in which only the upper 0.2-0.4 km was measured, and higher gradients are expected near the sea floor. Hill's value of 2.5 sec^{-1} also was measured over a depth of approximately 0.4 km. A possible alternate interpretation of his results, allowing that the closer points represent reflections instead of refractions, could be made which would give a lower value of gradient. As shown in Figures 2 and 5, if there is a sub-bottom reflector at a depth less than that reached by the ray corresponding to the cusp, the reflection curve R_{II} associated with it will join smoothly to the R_S curve, and it would be difficult to distinguish between the two types of arrival.

IV NORMAL INCIDENCE REFLECTIONS

In addition to variable angle reflection and refraction measurements, a large number of normal incidence reflection measurements have been made. Only a small percentage of these has been published (Hersey and Ewing, 1949) and (Shor, 1959). Hersey and Ewing made certain classifications of reflection records on the basis of the character of the bottom reflection and of the number and type of sub-bottom reflections. They were able to show some correlation between reflection types and physiographic provinces for certain areas of the western North Atlantic. Shor's measurements in the Pacific covered the boundary between the present day clay deposition and carbonate deposition areas and showed markedly thicker sedimentary cover in the latter (southern) region. His results also showed gneater accumulation of material in valleys than on hills.

Sub-bottom reflections on standard echo-sounding records have been discussed by Hersey and Rutstein (1958), Heezen, Tharp and Ewing (1959), Worzel (1959), and Ewing, Luskin, Roberts and Hirshman (1960). These studies have shown that in many areas, penetration of the sediments to depths greater than 100 feet is achieved, even at the relatively high frequencies employed in the sounding equipment (12 kcs/sec). The sub-bottom echoes are often stronger than the bottom reflection. In certain areas, particularly on moderate topographic highs, these sub-bottom interfaces can be traced for many miles. As in Shor's results, the depth to the sub-bottom reflector is usually correlated with topography; the upper layer thins or pinches out on hills and thickens in valleys. In abyssal plains, the sub-bottom reflectors are usually not as continuous as on the rises, as might be expected when one considers that the plains are subjected to turbidity current deposition and the rises only to pelagic sedimentation.

Efforts to sample specific sub-bottom interfaces with coring apparatus have been made on two occasions. Worzel identified a prominent reflector in the southeastern Pacific as a widespread layer of white volcanic ash. Ewing et al (in press) found that strong reflectors on the outer rise of the Puerto Rico trench correlated with increases in rigidity, presumably associated with increased carbonate content.

Continuous profiling of sub-bottom reflections in shallow water areas has been described by Knott and Hersey (1956), Moore (1960) and Beckmann et al (1959). This development has made it possible to obtain detailed surveys of sediment structure to depths of thousands of feet in some areas and offers great advances in the study of marine stratigraphy.

A new technique, utilizing a Seismic Profiler, for making seismic reflection measurements in deep water has recently been developed and employed with great success. Sediment thickness of 12,000-15,000 feet was measured in water deeper than 2000 fathoms. Small charges, up to 1/2 pound T.N.T., were used for the sound source, fired on a two-minute schedule with the ship traveling at 7-8 knots. The receiver, an AX-58C or K-100 hydrophone, was slacked for each shot. The Seismic Profiler is a modified Times Facsimile drum recorder fitted with a slip clutch and a triggered release which provides time synchronization of random pulses. The record produced is similar to that of a standard echo sounder or one made by the various sub-bottom depth recorders used in shallow water areas. A paper describing the equipment and results is in press, (Ewing and Tirey).

V SUMMARY

Measurements of velocity gradients by variable range reflection studies give average values between 0.9 and 1.4 sec^{-1} for the upper 0.2 to 0.4 km of sediments in deep water areas. With the evidence cited previously for low velocities immediately below the water-sediment interface, we can consider that the velocity vs. depth relationship most common in deep water sediments is approximately described by a parabolic or exponential function in which the upper sediments have velocities equal to or slightly lower than that in the water. At greater depths in the sediments the gradient is apparently only 20-30 percent of that and, in fact, may become insignificant at depth in some areas. There is some indication that the high gradients persist to some depth at which de-watering is largely achieved. Below this depth compaction would be retarded, resulting in lower gradients. Such behavior would be consistent with observations of porosity vs. pressure of the kind summarized by Hamilton (1959) on the consolidation of sediments. Porosity decreases rapidly at first, then more slowly with increasing pressure and would be expected to result in the variable gradients of the type observed.

As shown by Nafe and Drake (1957), gradients in shallow water sediments generally do not vary as markedly with depth as do those in deep water. Reference is suggested to their 1957 paper for an analysis of the relationships between compressional and shear wave velocities, porosity, density and other physical properties. The evidence they presented for differences between the velocity-depth dependence in deep and shallow water was statistical and based on tabulations of reported velocities and depths. For accurate information on the depth variation of velocity in the uppermost portion of the sedimentary column more sensitive methods are required. Seismic records obtained by shooting and recording on the ocean bottom should provide the necessary additional precision.

Thicknesses of low-velocity sediments measured by reflection and refraction techniques range from an average of 1 km in the Atlantic to 1/2 km in the Pacific. As suggested by Hamilton (1959) in a comprehensive review of compaction and lithification processes and of measured velocities and thicknesses, there is good reason to believe that the underlying "layer 2", described by Raitt (1956), is also sedimentary material, presumably well consolidated or lithified. This interpretation has been cited by Hamilton and others to establish agreement between the amount of material estimated to have been eroded from continents and that found by seismic measurements in the oceans.

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TABLE I

Measurements	Latitude	Longitude	Water Depth (R_{IO})	Velocity Gradient K sec^{-1}
S1-98 & 99	52 - 25S	40 - 35W	5.0	0.45
H-8	23 - 32N	71 - 05W	7.2	1.0
H-11	26 - 10N	75 - 02W	6.1	0.9
H-19	33 - 20N	71 - 30W	7.0	1.4
H-20	33 - 05N	73 - 45W	6.5	1.4
H-15	36 - 15N	67 - 10W	6.5	1.5
H-33	38 - 05N	70 - 00W	5.1	1.0
H-4	21 - 26N	67 - 29W	6.8	1.2
0-55	35 - 15N	64 - 30W	6.7	1.0
MH-52	53 - 50N	18 - 40W	3.2	2.5

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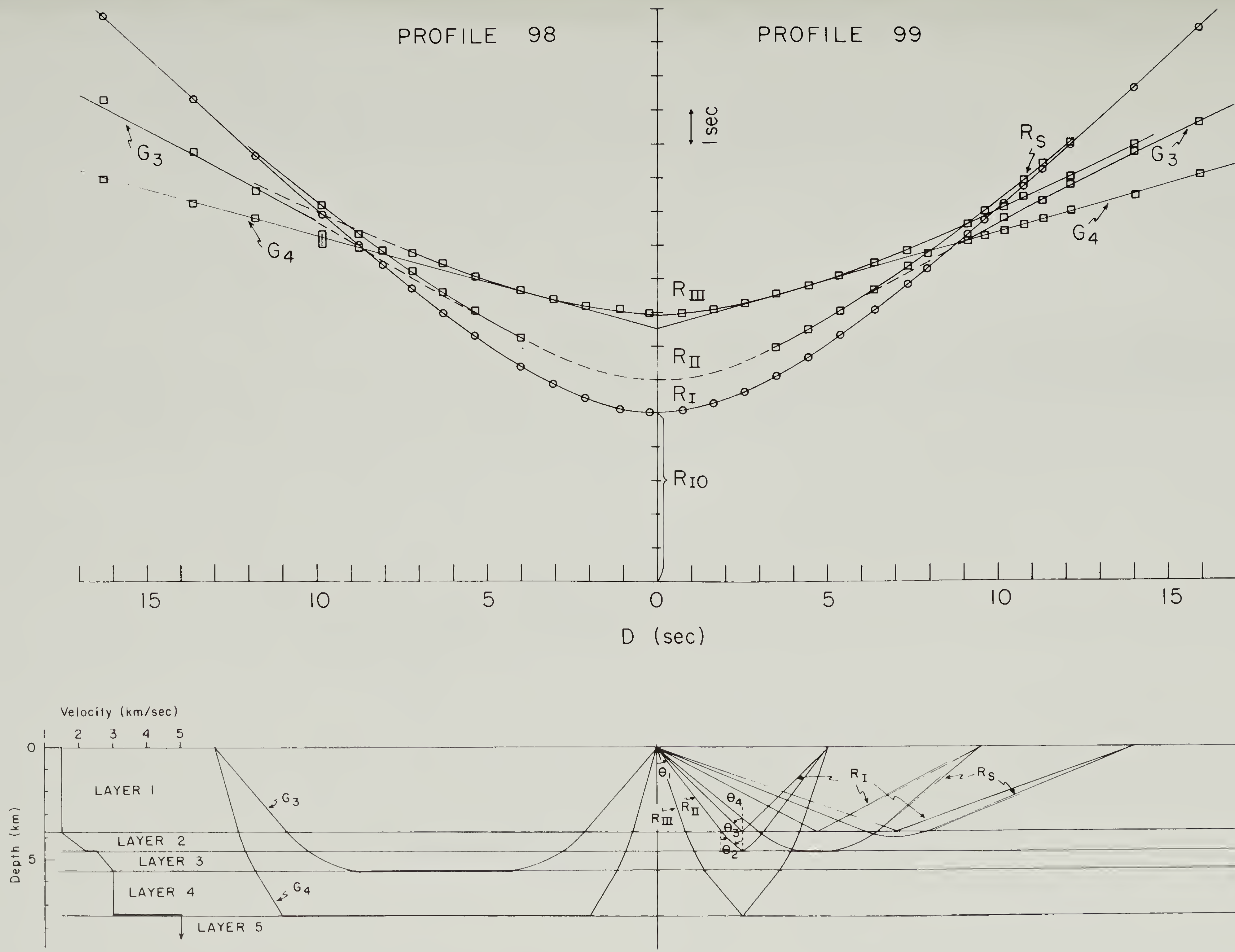
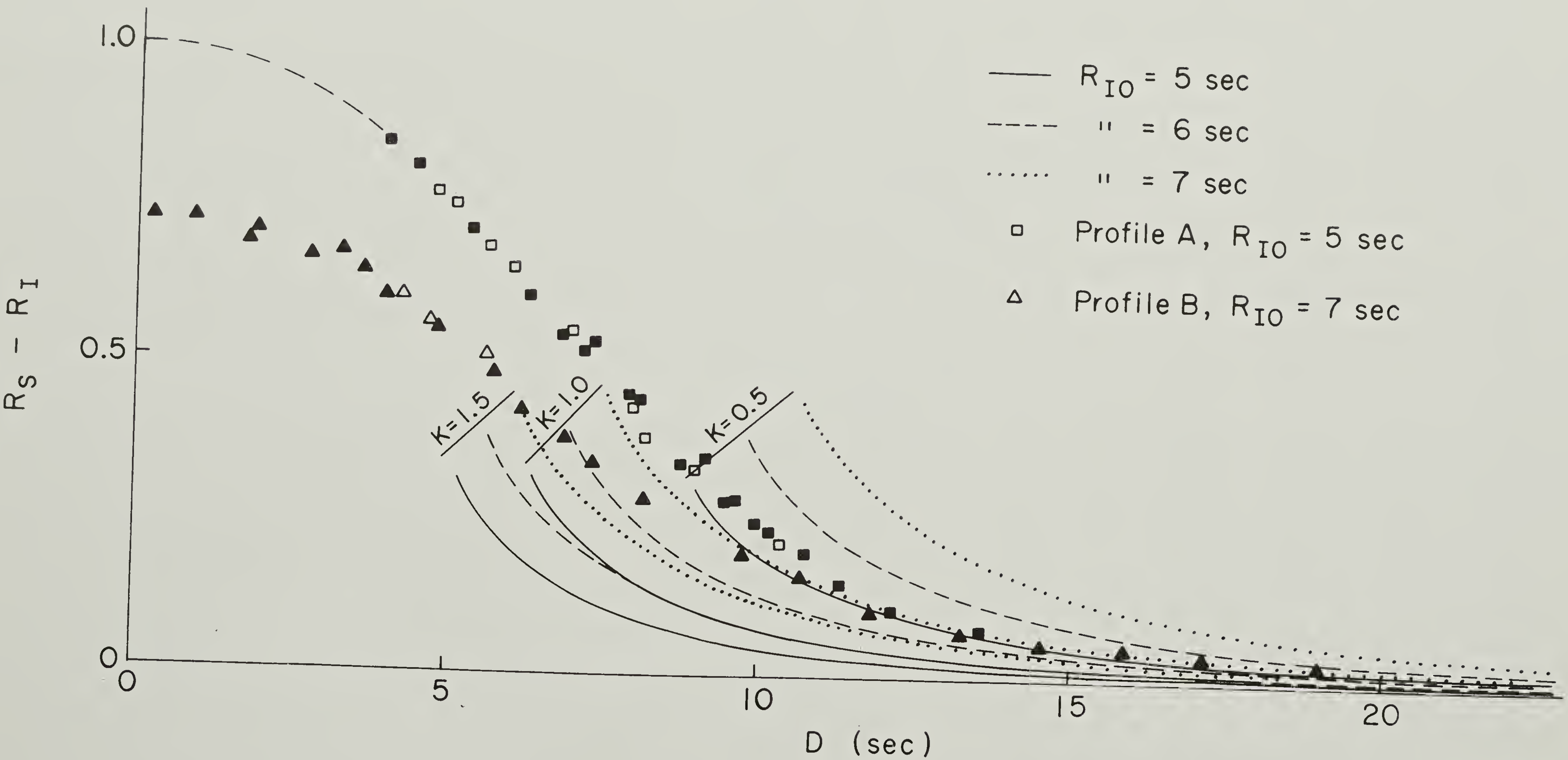


Figure 1. Time-distance graph, structure section and ray diagram for end-to-end profiles 98 and 99. Velocity vs. depth relationship shown at left of section. Ray paths at left of receiving position are for waves refracted at interfaces. At the right are ray paths for reflected waves and waves refracted by velocity gradient.



<u>Figure 2</u>. Theoretical curves for $R_S - R_I$ vs. D. Observational data from two profiles are plotted. Solid symbols represent arrivals observed in first order reflections and refractions; open symbols represent second order observations.

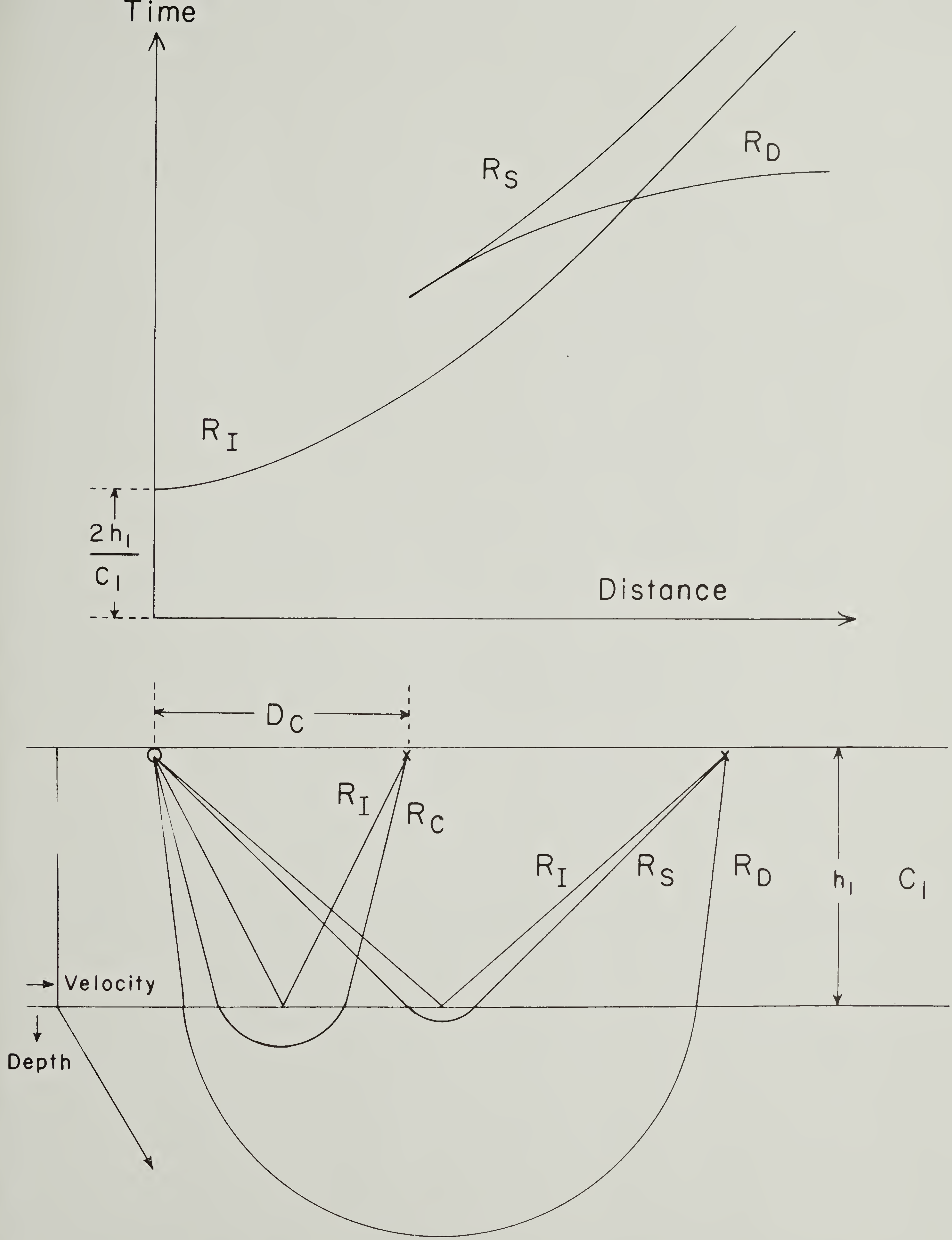
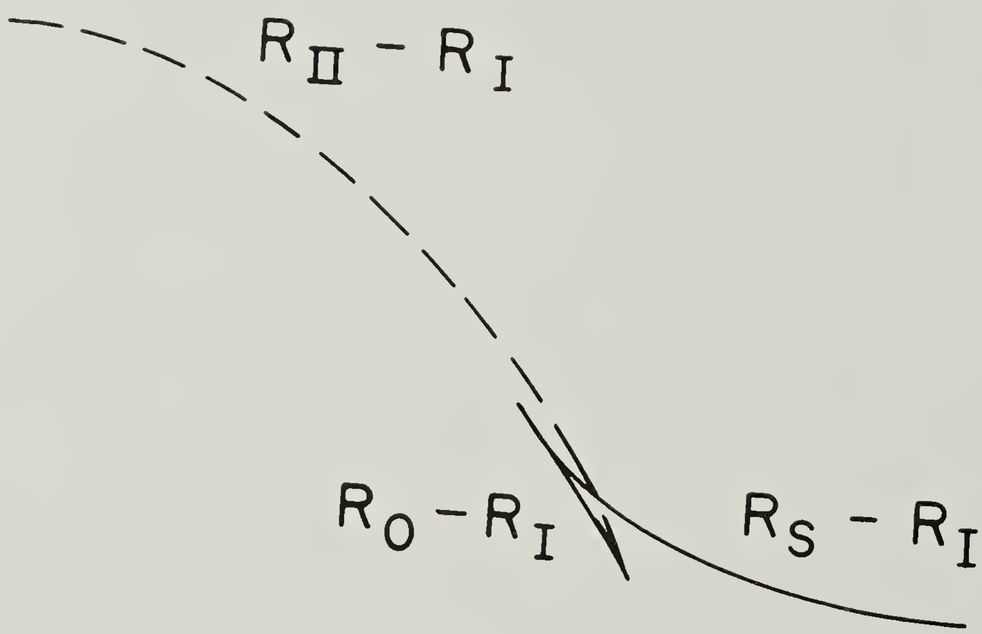
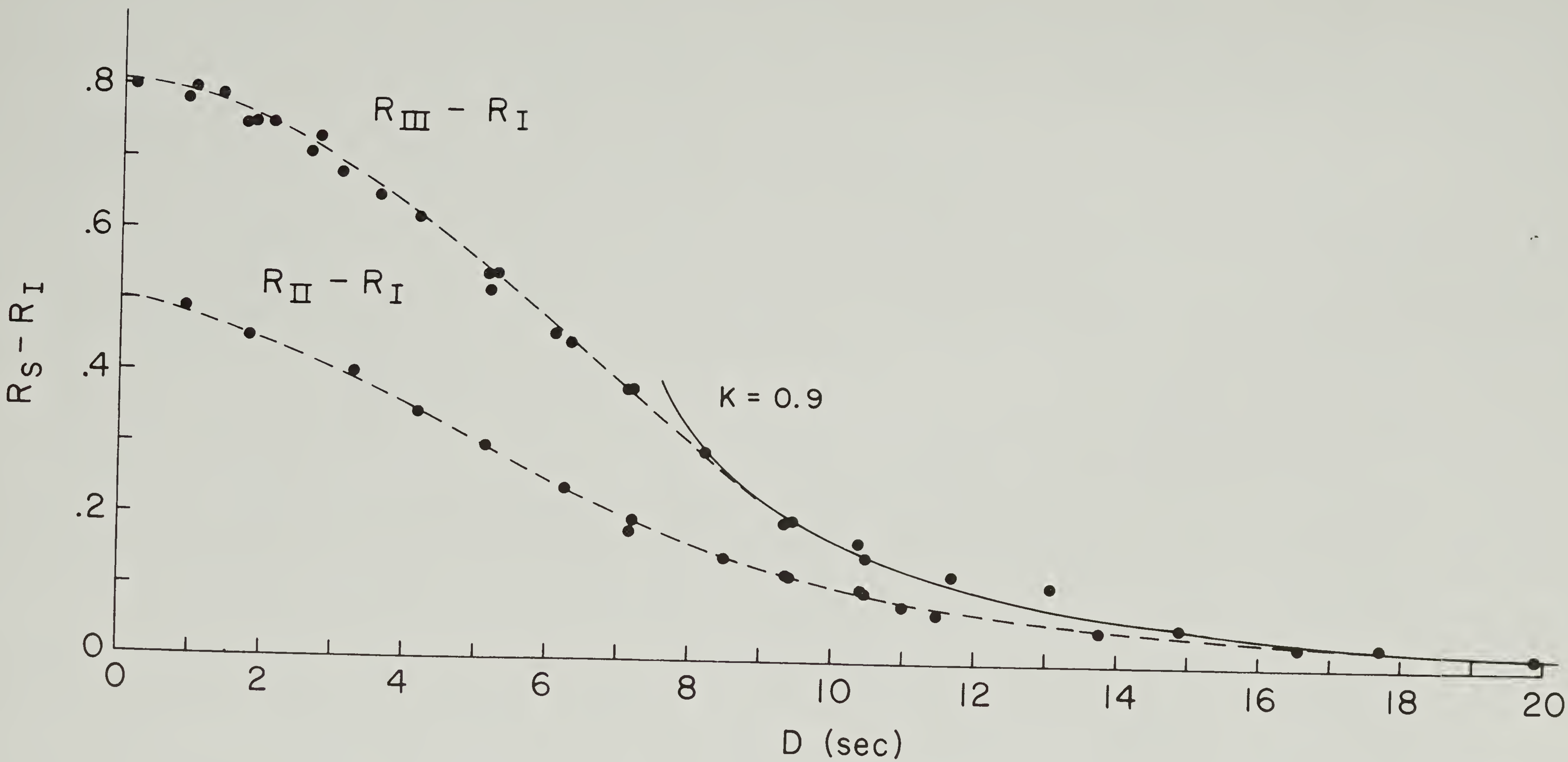


Figure 3. Ray diagram and time-distance graph showing effect of velocity gradient in thick sediments.



<u>Figure 4</u>. $R_S - R_I$ vs. D for case where 2-3 interface is deeper than depths penetrated by purely refracted ray.



<u>Figure 5.</u> $R_S - R_I$ vs. D for gradient 0.9 sec^{-1}, R_{IO} = 6 sec and two reflecting interfaces.

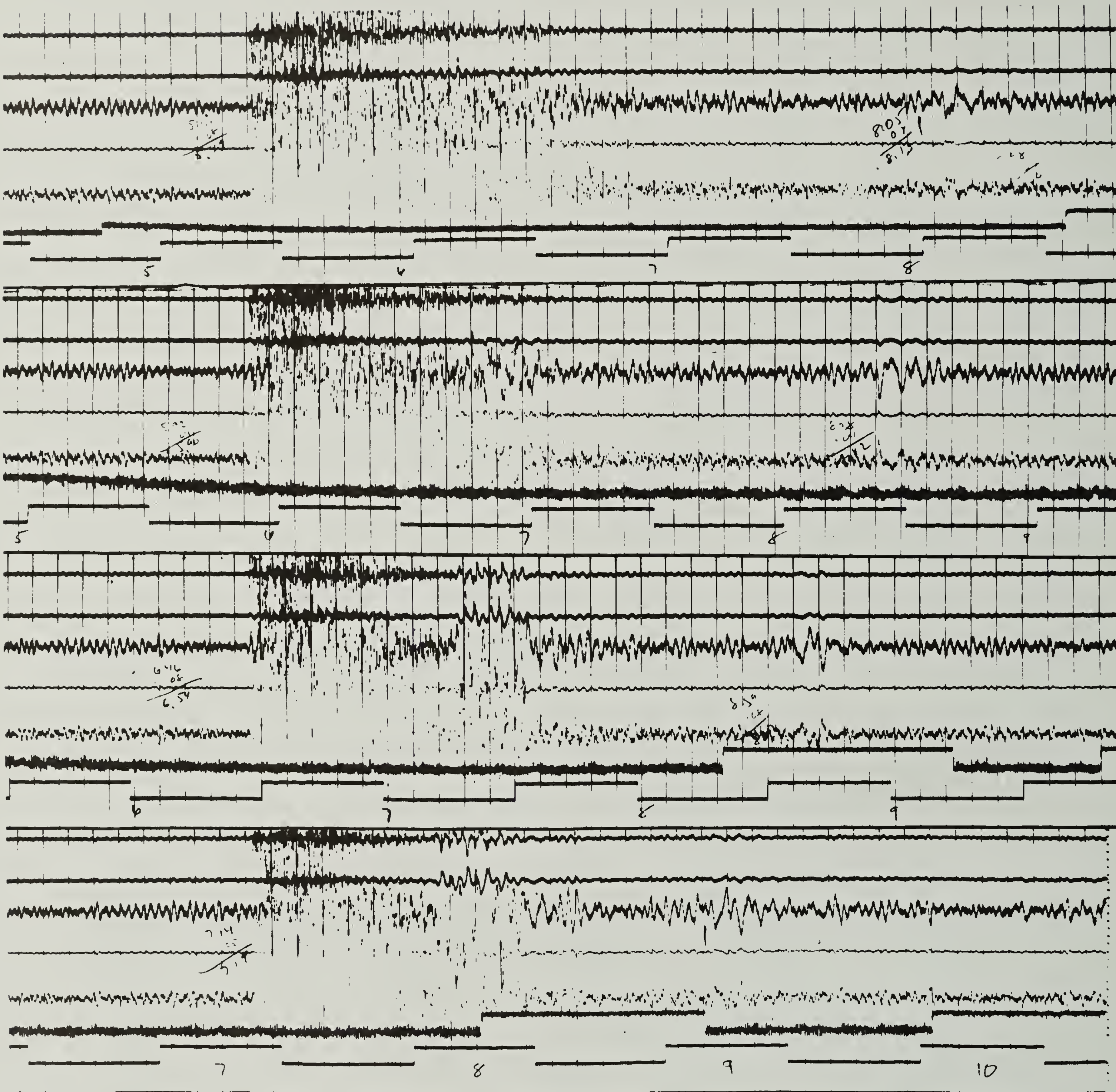


Figure 6. Seismic records showing critical angle reflections for R_{II}. R_{III} arrivals are from base of low-velocity sediments. Water depth = 3.75 km. Thickness of low-velocity sediments ≅ 4 km.